MARTIAN WATER

by

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ABSTRACT

The determination of the quantity of free water on Mars is approached through interpretation of the behavior of the polar caps under conditions of varying solar insolation. The vapor content of the atmosphere over the southern polar cap in spring is estimated to be 10^{-2} gm cm^{-2}. From this and the wind speeds expected on Mars a total quantity of water in the cap, and hence in circulation over the whole planet, is estimated to be 10^{16} gm. This is compatable with amounts observed spectroscopically to date. This amount of water serves to explain the dawn haze, and it is suggested that most of the atmospheric water condenses out at night and vaporizes again each day. It is also pointed out that since the humidity directly over the evaporating polar cap is essentially 100 %, and the temperatures are above 273° at the edge of the cap, that liquid water can be expected there. The calculations of the amount of water on Mars by Lebedinskii and Salova, made assuming zero humidity over the cap, are shown to be wrong by several orders of magnitude.

INTRODUCTION

For many years Mars was not a respectable topic of
serious study. The popularization of the canal theory and the
assertion that Mars was the abode of intelligent life by Lowell
and others at the beginning of this century arroused widespread
public interest in the planet but discouraged astronomers from
entering an arena so filled with misconceptions and controversy.
The slow but sure demise of the canals, and the interest now
being aroused at prospects of investigating the planets directly
from space probes, have again brought our nearer astronomical
neighbors into prominence.

This paper is a product of the renewed interest in planetary
astronomy. It does not present any new data or develop any
new theory of the Martian atmosphere. Rather, it is an attempt
to apply some rather well known principles from astronomy and
meteorology to the Martian atmosphere and polar caps. That
this approach is worthwhile is demonstrated by the fact that
some authors have constructed theories and deduced the amount of
water on Mars in contradiction to these principles. I feel that
my results for the amount of water on Mars are as accurate as
any published, in spite of their great intrinsic uncertainty.

My primary references have been the three books by
de Vaucouleurs [1954], Kuiper and Middlehurst [1961], and
Kellogg and Sagan [1962]. I have used the meteorological texts
by Hess [1959] and Fleagle and Businger [1963], and my main sources
for models of the Martian atmosphere have been Goody [1957]
and Ohring, Brooks, and Marino [1964]. Specific references are
given in the text.

The question of the presence of water on Mars can be
approached in three ways: First we may ask whether, from our
knowledge of the past history and present nature of the planet,
we can expect water to be present. Second, we can inquire
whether there is indirect evidence, in the form of surface

and atmospheric phenomena, for the existence of water on the planet.
And finally, we can endeavor to find direct evidence for water,
either spectroscopically, or by _in situ_ experiments carried
out on the Martian surface. This paper takes the second approach,
although the next section discusses the results of the first
approach,and the direct evidence, which is not large, is
reviewed in the last section.

SHOULD THERE BE WATER ON MARS?

Assigning an a priori probability to the existence of water
on the Martian surface would require a detailed knowledge of the
origin of planets and their evolutionary history. This much is
clear: hydrogen and oxygen are abundant elements and their most
stable compound, water, may be expected to have a relatively high
cosmic abundance where temperatures are low enough to avoid
decomposition. Argument by analogy with Earth is always risky,
but it is notable that terrestrial volcanoes constantly vent
water from the interior. The composition of Mars is not likely to
differ much from that of the Earth, so that it seems certain that
at some time Mars has had a considerable quantity of water, and it
is probable thatfor a significant fraction of its lifetime water
continued to be released into its atmosphere. Therefore we must ask
whether water will be able to exist for significant times on the
Martian surface or in its atmosphere.

Of great significance for all that follows is a knowledge of
the temperature of the Martian surface. All minerals which might
make up the crust of the planet are such poor conductors of heat
that we can neglect energy transport between the surface and the
interior. Thus the only source of energy is solar heat. For
the moment we will consider only two temperatures of the planet:
the equilibrium isothermal temperature and the equilibrium
temperature of the sub-solar point. To obtain the isothermal
temperature we equate energy received to energy radiated:

$$\frac{H_o}{r^2} (\pi R^2)(1 - A) = (4\pi R^2)\epsilon \sigma T^4$$

where H_o is the solar constant, r is the radius vector from the
sun to Mars expressed in AU, R is the radius of Mars, ϵ is the emissivity,
T is the desired temperature, and A is the albedo. The emissivity
will be near unity in the infra-red where the planet radiates;
therefore we can set $\epsilon =1$. The albedo, defined as the ratio of
total energy reflected or scattered back to the total energy
incident on the planet, is the least well known of the above

quantities. To evaluate it properly it is necessary to observe, at
all wavelengths and all phases, the intensity of the scattered
radiation. Most of the relevant regions of the spectrum are
available, but the maximum phase angle of Mars is only 47°, so that
a part of the phase function must remain unknown. A reasonable value
for A is that given by Kuiper [1952] and by Slipher [1962] of 0.15.
With these numbers we obtain a temperature of $T = 227^{\circ}K$. The
formula for the sub-solar point temperature will be identical to
the above except that the factor 4, which gave the difference in
receiving and radiating areas, will be omitted. The temperature
is then $\sqrt{2}$ times the above, or $311^{\circ}K$.

These calculations indicate a generally rigorous climate
on Mars, one in which water, if present, will usually be frozen
solid. It is interesting to see how authors a half century
ago used similar arguments to reach the opposite conclusion.
(i.e., Lowell, Philosophical Magazine, July 1907). Since the
value of the solar constant was not well known, a comparison
was made between the temperatures of Earth and Mars. It is clear
that the ratio will depend on the square root of the relative
distance and on the fourth root of the relative values of $(1 - A)$.
Lowell took an albedo for Earth of o.75 and of Mars of 0.27, from
which he concluded that the Martian temperature was 1.05 times
that of the Earth, indicating a tropical climate. His primary
error was, of course, in choosing too high an albedo for the
Earth. But if we use a more realistic albedo, such as 0.39, we
find that the temperature for the Earth as computed from the radiation
balance formula on the preceeding page is only $246^{\circ}K$, considerably
below what our experience would lead us to expect. The fact is
that 246° corresponds to an altitude of about 10 km above the
surface, where optical depth of unity in the infra-red occurs.
So we must be careful in the case of Mars, as well as Earth,
in going from the average temperature to the surface temperatures.
This topic will be treated in more detail in the next section,
but for the present we can conclude that Mars is colder than Earth.

If there were initially large quantities of water on the surface of Mars an equilibrium between the water (or ice) and a saturated vapor above it would establish itself. At 230°K the vapor pressure of water is 10^2 dyne cm^{-2}, or o.1 mb. This indicates a mass of water vapor in the atmosphere of about 0.2 gm cm^{-2}, for a saturated atmosphere; at higher temperatures this mass is much greater. As in the terrestrial case, condensation would produce clouds and precipitation. If Mars were once such a cloud-shrouded and ice-covered planet, would it long remain that way? One direction events would probably take is that the surface, blanketed by clouds and water vapor, would warm up, transforming the sterile ice sheets into teeming seas. Indeed, if there is or ever has been life on Mars, it must have arisen under such conditions. This fertile environment may have been short-lived, however, if the water vapor were able to escape into space.

The theory of escape of atmospheric gasses is extensively treated by a number of authors, and is specifically applied to Mars by Kellogg and Sagan [1962]. The conclusions depend critically on the temperature of the base of the exosphere, the effective layer from which escape takes place, and this temperature is not known with much accuracy. These authors conclude that the temperature in the critical layer is about 900°K, and from this they conclude that water vapor, once it is dissociated by solar ultraviolet radiation, will rapidly escape, so that the planet can not now retain any of its original atmospheric water. If there is water on Mars, then, there must be a source to make up for what escapes.

On Earth the primary source of water is volcanic escape from the interior. On Mars, there is no evidence for or against vulcanism, but there is another possible source suggested by Whipple. Mars is in a region of much higher meteoric density than Earth, and Whipple estimates that 10^5 tons of stony meteors may strike the planet per day. This is 3 X 10^{13} gm/year; if one part in ten thousand is water, this supplies 10^9 gm of water per year. If, as seems likely, the characteristic escape time of Martian water is at least 10^6 years, this leads to at least 10^{15} gm of water in the atmosphere in an equilibrium state.

This approach has really gained us nothing more than a plausibility argument for free Martian water. Indeed, this must be, for so many of the relevant numbers are unknown, even to order-of-magnitude accuracy. It is necessary to consider what observational data lead us to believe that water is present on the planet, and to ask what quantitative interpretation can be placed on these data.

THE TELESCOPIC VIEWPOINT

To the observer with a small telescope the most striking feature of Mars is its polar cap of shining white. The analogy with the terrestrial ice caps is inescapable, and the seasonal waxing and waning of the Martian caps, first noted by Sir William Herschel, further supports this analogy. It is clear that some white material is deposited at the winter pole and that, as spring brings increased solar radiation to the polar region, these "night-caps of the planet's winter sleep" progressively disappear, leaving only a small cap in the northern summer and quite anihilating the southern cap during its relatively warm summer. Two substances which might present such a cycle come to mind: CO_2 and H_2O. But the temperatures on Mars are too high for "dry ice", leaving us with true ice, a conclusion which has been amply confirmed from reflection spectra [Kuiper, 1952] and polarimetric observations [Dollfus 1961]. The polar caps are ice; there is no doubt about it. The only alternative suggested in recent years is nitrogen dioxide [Kiess, Karrer, and Kiess 1960], but this hypothesis meets enormous difficulties, both observational and theoretical [Sagan 1964], and deserves no serious consideration here.

The polar caps are the most striking feature of the planet, and their evolution is intimately connected with all of the enigmatic changes which are seen on the planet. As Lowell has so well put it [1906, p 37]:

It is not a matter of hazard that the most evident of all the planet's markings should also be the most fundamental, the fountainhead from which everything else flows. It is of the essence of the planet's condition and furnishes the key to its comprehension. [The caps'] size first riveted man's attention, and then attention to them disclosed that most vital of the characteristics of the planet's surface: change.

With the coming of spring the winter cap retreats, and as it does so, changes take place which have intrigued astronomers and aroused widespread public interest for nearly a century. Most of the changes are manifestations of what has been called

the "wave of darkening". In rough progression downward from the retreating ice the dark markings become darker and their outlines sharper; they remain this way for several weeks and then gradually return to their former dull grey. The wave of darkening moves at a mean speed of about 35 km/day, progressing to the equator and spilling over into the opposite hemisphere, where it finally fades out. The southern cap is larger and retreats faster, and the southern wave of darkening is more conspicuous than its northern counterpart. If the wave of darkening is extrapolated at constant speed to the opposite pole, it is found that its time of arrival corresponds approximately to the time when clouds form over the pole before it is lost to view in the long polar night, from which it will emerge sparkling into the clear spring light half a year later. This never ending progression: cap to wave of darkening and back to cap, is the dominant pattern against which a myriad of smaller variations take place.

It is difficult to escape the conclusion that the substance of the polar caps is being transported from one hemisphere to the other with the alternation of the Martian seasons, and that in its passage it strongly effects the surface of the dark regions while leaving the lighter deserts unchanged. Since the caps are ice, it must be some form of water which is transported. And anyone who has seen the deserts of Earth bloom into a riot of life after a rain must ask if the Martian water is not producing a similar effect in the arid plains of that planet. The possibility that life exists on Mars, and the necessity of water for life, lend especial interest to an enquiry into the quantity of water available to these hypothetical organisms.

During their rapid retreat, the caps are usually surrounded by a dark collar some 100 km or more wide. This collar follows the shrinking cap and seems to represent a strong darkening of the surface in the wake of the ice. Lowell saw this collar as deep blue in color and assumed that it consisted of lakes of water ringing the ice cap. Indeed, he used this existence of

liquid runoff from the cap as proof that the caps were of ice and not frozen CO_2, which would sublimate without passing through the liquid phase. These supposed lakes are at such high Martian latitudes that the hypothesis could not be checked by looking for specular solar reflection from the liquid surface, and for many years it tended to be generally accepted. More recently, polarimetric measurements by Dollfus have led him to deny the possibility that the dark collar is due to liquid water, since the granular nature of the surface is not changed by the darkening of the collar; but admittedly the observations are at the limit of his equipment. The explaination of the dark collar is not, however, so easily fielded to the exobiologists as is the wave of darkening, and this is a subject to which I will return.

Ephemeral light patches are seen obscuring the surface features from time to time, and these almost certainly are atmospheric phenomina. The "yellow clouds" are not directly related to the problem of water, but their motions do give evidence of wind velocities up to 30 km/hour. Of more interest are "white clouds" which are common near the limbs and at the poles and are seen occasionally over the central part of the disk. Polarimetric studies [Dollfus 1957] support the view that these clouds are ice crystals, similar to terrestrial cirrus. There is also a general obscuration of the planetary surface in the blue and ultra-violet, the so-called "blue haze", which occasionally "clears" over the whole planet in times of no more than a day; this haze may also be due to atmospheric water. Observations of all these phenomena are difficult to make, and little is established beyond their existence. Still, their presence does set some constraints on the quantity and state of water on the planet.

Telescopic observation establishes that there is ice on Mars, at least in the polar caps, and it hints at the presence of cirrus clouds and surface wetting. Most challengingly, it demonstrates striking seasonal changes which seem to be triggered by release of the polar water. But before we can try to explain these observations, we must look more closely at the climatic conditions which prevail on Mars.

MARTIAN METEOROLOGY

Only when a planet posseses an atmosphere can we speak
of it as experiencing weather. An unshielded surface is subject
only to the heat of the sun alternating with the cold of night.
But within a sphere of air, the fundamental alternation of
radiative heating and cooling is tempered in a myriad ways, as
is so well demonstrated by the terrestrial weather, which remains
so varied and unpredictable that it is a never-ending subject of
conversation. Mars has an atmosphere; therefore its weather must
be tremendously complicated. Indeed, if we cannot understand
and predict our own weather, how hopeless *must* be its study on
a globe never more than a few seconds of arc in radius. And yet,
some understanding of the principles of Martian weather and climate
is necessary if we are to proceed in our discussion of water on
that planet.

We have shown that the equilibrium temperature of Mars,
assuming it to be isothermal, is 227°K. We must now consider the
variation of surface temperature with both time and latitude.
The variation of solar heating as a function of these two variables
is the primary control of the surface temperature and of the
energy sources which drive atmospheric motions. In considering
surface temperatures the effect of the atmosphere can be neglected,
to a good first approximation, since the atmosphere is largely
transparent to solar radiation. The atmosphere will certainly
be important in keeping the surface temperature from dropping
to absolute zero at night. On the daytime side, it will act to
reduce the extremes computed from solar heating, but it will
also slightly raise the temperature by the greenhouse effect of
CO_2, so that the net effect will be little change from the
maximum temperatures we will compute by neglecting the atmosphere
entirely.

Before proceeding it is appropriate to banish all worries
over conduction of heat to or from the interior. It is simple
to estimate the order of magnitude of the fluxes conducted. The

heat flow equation is:

$$\frac{\partial T}{\partial t} = K_s \frac{\partial^2 T}{\partial z^2}$$

where K_s is the thermal diffusivity and z is measured downward
from the surface. Replacing differentials by finite differences
and solving for Δz, the skin depth, gives:

$$\Delta z = (K_s \Delta t)^{1/2}$$

Let us consider a characteristic skin depth for Mars. A reasonable
value for K_s is 10^{-3} cm^2 sec^{-1}, while for Δt we can take half a
Martian year, or 3×10^7 sec. This gives about 100 cm as the
depth over which temperature will change significantly from summer
to winter. To find the heat flux, we use the equation:

$$Flux = - K_o \frac{\partial T}{\partial z}$$

where K_o is the thermal conductivity and is about 10^{-3} cal cm^{-2}
sec^{-1} $^oK^{-1}$. The seasonal temperature range is no more than 100°;
using this and the skin depth found above gives a flux of the order
of 10^{-3} cal cm^{-2} sec^{-1}. This is down three orders of magnitude
from the solar flux incident at equatorial noon and can safely
be neglected in all of our considerations. The diurnal temperature
variations are as great as the seasonal ones, while the times are
two orders of magnitude less: thus the fluxes are one order of
magnitude greater. But this is still too little to effect the
rough calculations to be presented here.

The solar flux incident on the surface, or insolation, H,
is given by:

$$H = \frac{H_o}{r^2} \cos z$$

where z is the zenith angle of the sun. Assuming radiative
equilibrium, we have for the surface temperature:

$$T_o = \left[\frac{H_o}{\sigma r^2} (1 - A) \cos z \right]^{1/4} .$$

For any point on the surface z will have both a diurnal
and a seasonal variation. Only at the poles is the diurnal variation
absent, and here it is easy to find the insolation as a function
of time. Given insolation and albedo, we can find the equilibrium
temperature. For the polar caps an albedo of 0.5 is used, while
the value of 0.15 used before is taken as representative of the
rest of the planet. For these two albedos, and the aereographic
latitudes -60°, -75°, and -90°, I have computed the temperatures
at noon during the southern spring; these temperatures are shown
in Figure 1. The true temperatures are probably between the
two lines, since even where the albedo is only 0.50 there will
be some warming by advection from the nearby darker areas.

Figure 2, taken from Slipher [1962], shows the boundary of
the southern cap as it retreats during the spring. For the
latitudes and times corresponding to the edge of the cap, I have
computed the noon temperatures, and these are shown in Figure 3.
The corresponding temperatures during the northern spring will all
be lower by a factor of 0.91, due to the greater distance of
Mars from the sun at that time.

If the insolation is averaged from sunrise to sunset for
each latitude, this can be used to compute a sort of average
daytime temperature. This has been done by Sagan in an unpublished
work, assuming an albedo for the dark areas of 0.10, and his
temperature map is shown in Figure 4. It is clear from this
figure that the wave of darkening takes place at the times and
places that the temperature is a maximum; however, it is not
clear that there is a "wave" of darkening, as opposed to a
darkening simultaniously at all latitudes once the temperature
rises to about 300°. It is not really certain from present data
whether the darkening is due to polar water or not.

It would be unnecessary to resort to these rather idealized
theoretical temperatures if the observed temperatures of the planet
were not even more uncertain. Radio observations do no good,
for they cannot resolve the disk. Infrared radiometery has been
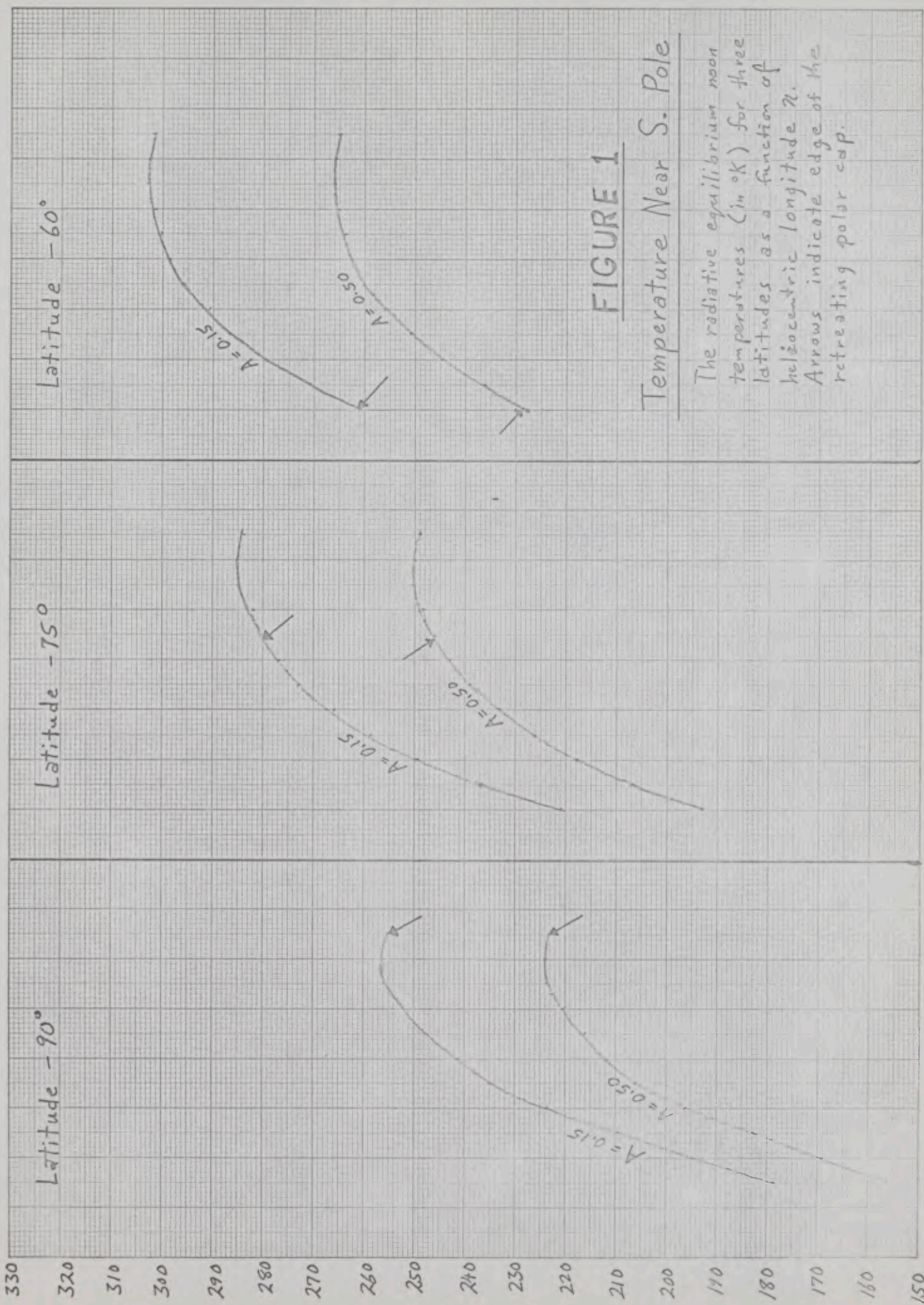
FIGURE 1

Temperature Near S. Pole

The radiative equilibrium noon temperatures (in °K) for three latitudes as a function of heliocentric longitude λ. Arrows indicate edge of the retreating polar cap.
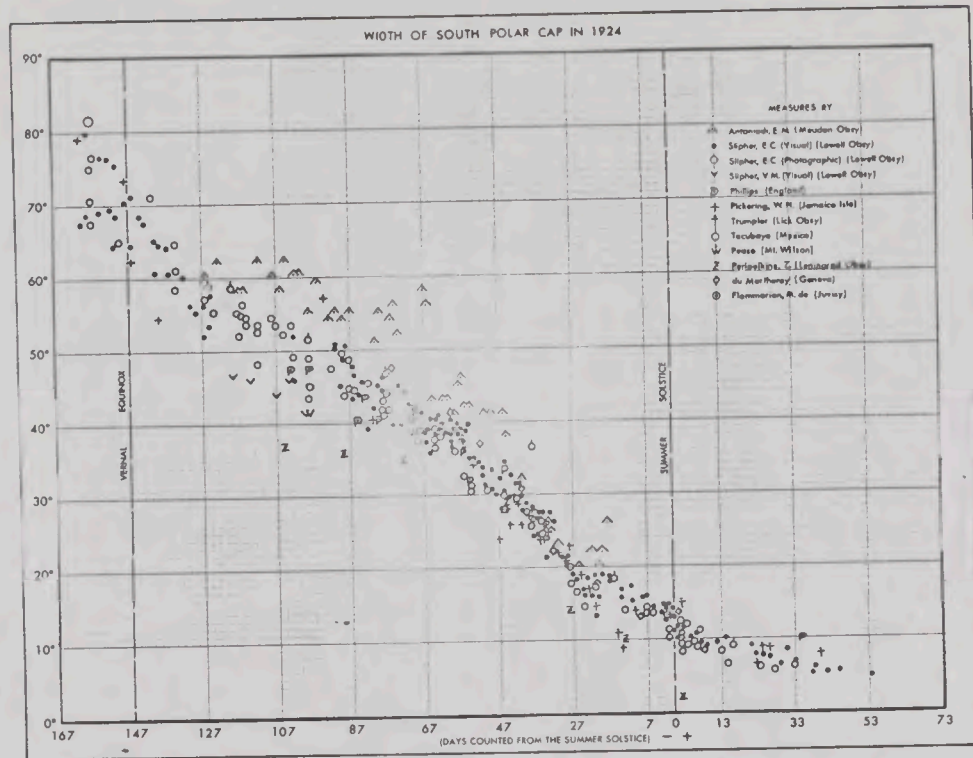
FIGURE 2

FIGURE 4. *This figure shows all the available observations (first mapped in 1925) of the size of the southern polar cap of Mars during its spring-summer recession at the favorable opposition (the nearest ever possible) in 1924. These observations embrace the Martian season from the vernal equinox to 54 days after the summer solstice of the planet's southern hemisphere. Without exception all the observations by more than a dozen observers working at nine different observatories, including a six-month series of photographs, reveal that the reported retardation in the melting of the cap in 1924 by Antoniadi was not observed elsewhere. Thus the observations indicate with the utmost clarity that the recession of the cap in 1924 was essentially normal.*

FIGURE 3

Noon Temperature of Edge of Cap

The radiative equilibrium temperature
at the edge of the South Polar Cap
during spring, given as a function
of the heliocentric longitude η,
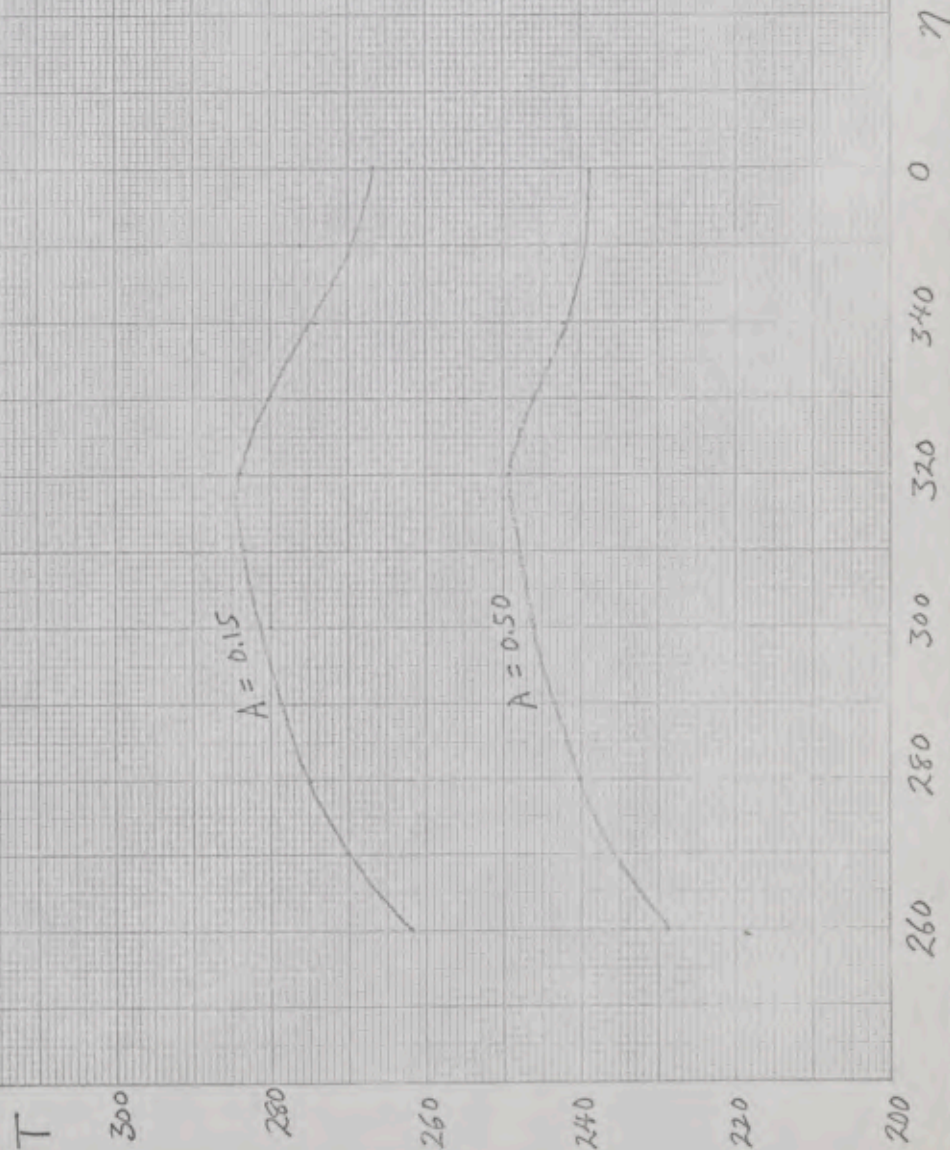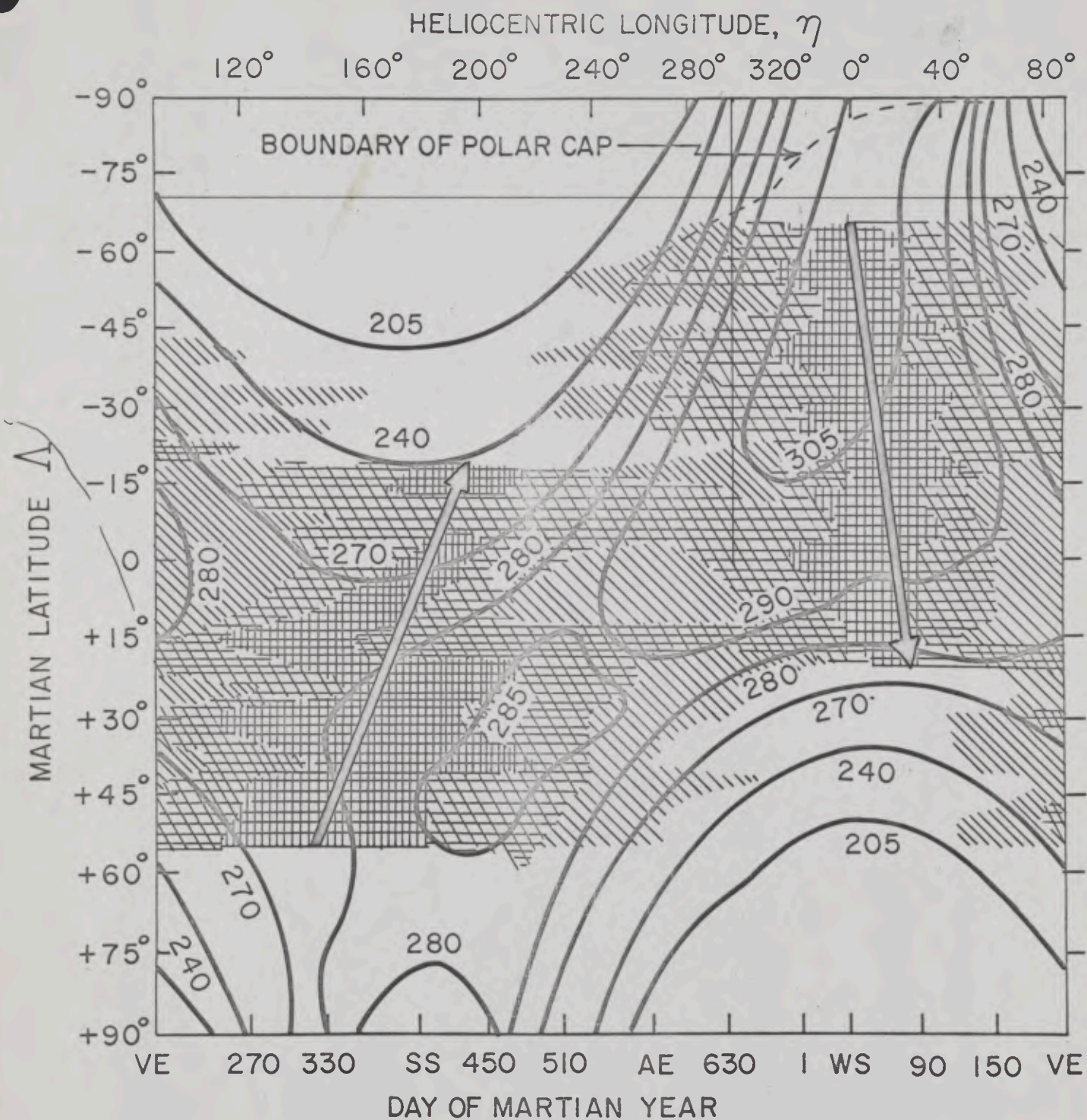computed for two values of the
albedo A.

FIGURE 4

applied for many years [see Pettit 1961], but the best that has
been achieved is a resolution of 1/8 of the disk and an accuracy
of $\pm 10^{\circ}$. The results are compatable with the theoretical temper-
atures given above, although they tend to be about 20° cooler.
The best results for the southern polar cap are by Gifford [1956],
who finds noon temperatures of less than 200° in winter, 230°
to 240° in spring, and 240° to 250° in summer. He states that
these are likely to be underestimates, although the magnitude of
the difference is unknown.

That Mars posseses an atmosphere is evident, but the deter-
mination of the composition and quantity of this atmosphere
is a problem not yet satisfactorily solved. The principal problem
is to find the surface pressure. Many ingenious, if rather
inaccurate, methods have been used to estimate this quantity
from the observed scattering, absorption, and polarization of light
passing through the atmosphere [see de Vaucouleurs 1954]. The
value for the surface pressure deduced from polarimetric studies
by Dollfus [1961] of 85 mb has long been the accepted value,
although it has frequently been pointed out [Goody, 1957] that
this assumed a dust and droplet free atmosphere, and that the
presence of these other scatterers would make the true pressure
lower than this estimate. Recently Kaplan, Münch, and Spinrad
[1964] have obtained spēctra of Mars in the near infra-red and
have used the pressure broadening of CO_2 bands to determine the
surface pressure. Their result is: $P = 25 \pm 10$ mb. Their
arguments for a value in this range seem quite convincing.
Kuiper is said to have obtained an unpublished pressure of about
10 mb, but this is probably a lower limit. We shall consider
25 mb to be the most probable pressure, but will note that values
from 10 to 50 mb are not excluded.

Given the surface pressure (and hence density, using the
perfect gas law), we can obtain the variation of these parameters
with altitude. In all cases the equation of hydrostatic
equilibrium will be closely obeyed:

$$\frac{dp}{dz} = -g\rho$$

where z is measured upward from the surface, ρ is the density, and g is the gravitational acceleration. Substituting into the above from the equation of state:

$$dp = -\frac{pg\mu}{RT}dz$$

Formally, the integration yields:

$$p = p_o \exp - \int_0^z \frac{gz\mu}{RT} dz$$

where p_o is the surface pressure. It is usually valid to neglect variation of g with z, and we shall make this assumption. The simplest solution is for an isothermal atmosphere; this gives the barometric equation:

$$p = p_o e^{-z/H}$$

where we define the scale height, H, as:

$$H = \frac{RT}{\mu g} = \frac{kT}{mg}$$

where m is the mass of a molecule. The physical significance of the scale height is as the total thickness the atmosphere would have if it had a constant density equal to the density at the surface. Even for cases where T varies with height, the concept of a local scale height is often useful.

In a real atmosphere isothermality is rather an artificial and unlikely requirement, although over small ranges of z the barometric formula is reasonably accurate. A much more realistic requirement is that all atmospheric changes be adiabatic. Since the atmosphere does not readily exchange heat with its surroundings, and since it is constantly in motion, the adiabatic assumption is a very good one, at least for the bottom few km of the atmosphere, where mixing takes place. From the first law of thermodynamics:

$$O = c_p \, dT - v \, dp$$

Combining with the perfect gas law:

$$O = \frac{dT}{T} - \frac{R}{\mu c_p} \frac{dp}{p}$$

This is integrated to give:

$$T = \text{const. } p^{\varkappa}$$

where $\varkappa$ is defined as:

$$\varkappa = \frac{R}{\mu c_p}$$

To find the variation of temperature with altitude, we differentiate the adiabatic equation logarithmically with respect to z:

$$\frac{1}{T} \frac{dT}{dz} = \frac{\varkappa}{p} \frac{dp}{dz}$$

Substituting from the hydrostatic equation and the equation of state yields:

$$\frac{dT}{dz} = - \frac{g}{c_p}$$

The lapse rate is defined as the negative of this gradient:

$$\Gamma_d = \frac{g}{c_p} \quad -$$

where the subscript d refers to the fact that we have been considering only dry air obeying the perfect gas law. The variation of pressure with altitude is found by noting that the lapse rate is constant, so that we may write for the temperature:

$$T = T_o - \Gamma_d \, z$$

Then we have:

$$p = p_o \exp - \frac{g}{R} \int_0^z \frac{dz}{T_o - \Gamma_d \, z}$$

$$p = p_o \left(\frac{T}{T_o} \right)^{\frac{g\mu}{R\Gamma_d}} \quad .$$

On Earth the atmosphere is greatly affected by processes involving phase changes of water; since the energies involved are large, the departures from dry adiabatic behavior are large. However, for Mars we can use the adiabatic equations with confidence in most cases, since the water there is manifestly much less than on Earth. From the outset we can be sure that no hurricanes or thunderstorms complicate the weather on that world.

Kaplan, Munch, and Spinrad [1964] have determined the CO_2 content of the Martian atmosphere to be 55 + 20 m atm; presumably most of the rest of the atmosphere is N_2, since no O_2 has been detected. They indicate the most probable composition to be: 19 mb N_2, 2 mb A, and 4 mb CO_2. These are close enough to terrestrial values to let us transfer values for such terms as μ and c_p directly from terrestrial experience. The primary differences arise from the lower gravity on Mars; thus the scale height is about three times greater than on Earth and the lapse rate about three times less. This leads to the curious fact that at very high altitudes the Martian pressure and density are greater than the terrestrial at the same altitudes, while the gentler temperature gradient just keeps pace with the gentler fall-off of pressure with altitude.

In summary, the following table of the thermodynamic constants of the Martian atmosphere is adapted from Ohring, Brooks, and Marino [1964], using the composition suggested by Kaplan et al.:

Molecular Weight	μ	30.52
Specific Gas Constant	R/μ	2.73×10^6 erg gm^{-1} deg^{-1}
Specific Heat at Const. p	c_p	9.65×10^7 erg gm^{-1} deg^{-1}
Poisson Constant	κ	0.282
Surface Gravity	g	376 cm sec^{-2}
Scale Height at $0^{\circ}C$	H	19.8 km
Dry Adiabatic Lapse Rate	Γ_d	3.9 $^{\circ}$/km

The structure of the lower atmosphere of Mars must be qualitatively similar to that of the Earth, consisting of a convective troposphere below a nearly isothermal stratosphere. The temperature of the stratosphere can be computed on the assumption that it is transparent in the visible and is sufficiently opaque in the infrared to be in radiative equilibrium with the surface. This assumption, while crude, does give nearly correct values for Earth and presumably also does so for Mars. Under this assumption the stratosphere receives the blackbody radiation of temperature T_o from one direction (down) but radiates at its own temperature T_s in two directions, so that:

$$T_s = \frac{1}{\sqrt[4]{2}} \, T_o$$

The value of the lapse rate in the troposphere is known, so that we can find h, the altitude of the tropopause (base of the stratosphere). The actual height may be somewhat higher than the value computed in this way, since the tropause is the coldest part of the atmosphere and the stratospheric temperature we have found may apply to somewhat higher and warmer layers. The following table shows the values given by this simple theory for the height, temperature, pressure, local scale height , and saturation vapor pressure and mixing ratio of water, all at the tropopause, and computed for three values of the surface temperature:

T_o [°K]	$T(h)$ [°K]	h [km]	$p(h)$ [mb]	$H(h)$ [km]	$p_w(h)$ [mb]	w_ε
300	252	12	14	18	7.7×10^{-1}	3×10^{-2}
250	210	10	14	15	8.2×10^{-3}	3×10^{-4}
200	168	8	14	12	$< 10^{-4}$	$< 10^{-6}$

With this crude model atmosphere we can find the altitudes at which given mixing ratios of water will condense into clouds, and we can estimate the thermal and eddy diffusion rates for vaporizing water. We can also go on to consider the general circulation of the atmosphere and estimate the winds available

to transport water vapor on Mars.

Winds are driven by differential solar heating and are directed
by Coriolis forces generated on the surface of a slowly rotating
planet. In both respects Mars, whose polar inclination and
rotation period are practically identical to Earth's, yields
results closely parallel to the terrestrial situation.

The net effect of differential solar heating and radiative
cooling is to create a heat source at the equator and sink at the
poles. In this situation two modes of atmospheric circulation
are possible: the symmetric regime and the wave regime. Yale
Mintz [1961] has discussed the criteria for stability of these
two flow patterns and had applied the simplified theory to the
Martian atmosphere.

In the symmetric regime the warm air rises near the equator
and moves toward the pole, where it descends, while cool air moves
near the surface from arctic to tropics. The motion is not,
however, along the meridians; the rotation of the planet causes
the winds to spiral around the pole. In the wave regime this
simple motion is unable to transport the necessary energy poleward,
and it breaks down into a number of large amplitude horizontal
waves and eddies. This complex pattern is typical of terrestrial
circulation.

Mintz shows that the regime which will predominate is a
function of the poleward temperature gradient. Although the
Martian geometry is almost identical to the Earth's, and hence
the differential solar heating the same, the net differential
heating is quite different. This is because the effective
terrestrial radiator is water vapor in the troposphere, and the
layer of optical depth unity in the infra-red is practically
isothermal, so that a relatively large amount of energy is radiated
from the polar regions even though the surface temperatures are
much lower than near the equator. On Mars the bulk of the radiation
comes from the surface; this is why we have been able to compute
surface temperatures from simple radiation balance arguments.

Normally the poles are colder, and they radiate less, thus conserving heat and decreasing the net differential heating relative to the Earth. Of course, our radiation balance temperatures assumed that the net differential heating was zero, with radiative equilibrium prevailing at each point and no heat left over to be transported by the atmosphere.

When account is taken of the effect of the atmosphere on solar and surface radiation the latitudinal temperature gradient can be estimated; Mintz finds it is near the critical value which separates the wave and symmetric regimes. This it is not possible, with present knowledge, to say which circulation pattern prevails; in fact, different regimes may apply at different seasons. Our ignorance precludes estimates of wind velocities or of the degree of interhemispheric circulation. We are simply warned that terrestrial analogies are dangerous, in spite of the apparent similarity of Earth and Mars.

Another approach to Martian meteorology involves an application of the geostrophic and thermal wind equations. (These are topics treated in standard texts on meteorology; see, for instance, . Hess 1959). Geostrophic flow occurs when the motion of air is such as to balance a pressure gradient force by the Coriolis force at each point of its motion. In equilibrium the motion will be parallel to the isobars and at constant speed. For a particle of mass m in a uniformly rotating Cartesian frame the equations of motion are:

$$\ddot{x} = 2\,\Omega\,(v \sin\phi - w \cos\phi) + F_x/m$$
$$\ddot{y} = -2\Omega u \sin\phi + F_y/m$$
$$\ddot{z} = 2\,\Omega\,u \cos\phi - g + F_z/m$$

where Ω is the angular rotation rate, ϕ is the latitude, and the last terms on the right represent all external forces. The centrifugal force is combined into the gravity g. The terms $2\,\Omega \sin\phi$ are usually defined as the Coriolis parameter f. Pressure forces per unit mass of fluid are given by:

$$F_x = - \frac{1}{\rho} \frac{\partial p}{\partial x}$$

In the case of geostrophic flow all accelerations are zero, and we do not permit vertical motions. The equations of motion then reduce to the geostrophic equations:

$$f \, v = \frac{1}{\rho} \frac{\partial p}{\partial x}$$

$$f \, u = - \frac{1}{\rho} \frac{\partial p}{\partial y}$$

$$g = - \frac{1}{\rho} \frac{\partial p}{\partial z} + 2 \, \Omega \, u \cos \phi$$

If we neglect the small Coriolis term, this last is simply the hydrostatic equation. On Mars f has about the same value as on Earth, the density is much lower, and the pressure gradients are much less. These last two effects compensate for each other, and so we are led to expect comparable geostrophic winds on Mars and Earth.

We now relate the vertical shear of the geostrophic wind to the horizontal pressure gradient. The geostrophic wind will obey, in addition to the three equations given above, the equation of state:

$$\rho = \frac{p \mu}{RT}$$

This gives four equations in the five variables u, v, p, ρ, and T. We cannot solve, but we obtain an interesting inter-relationship. Eliminating the density from the geostrophic equations we have:

$$\frac{f \, v}{T} = \frac{R}{\mu} \frac{\partial \ln p}{\partial x}$$

$$\frac{f \, u}{T} = - \frac{R}{\mu} \frac{\partial \ln p}{\partial y}$$

$$\frac{g}{T} = - \frac{R}{\mu} \frac{\partial \ln p}{\partial z}$$

We eliminate the pressure, reducing the above equations to two by cross differentiation:

$$\frac{\partial}{\partial z}\left(\frac{fv}{T}\right) = -\frac{\partial}{\partial x}\left(\frac{g}{T}\right)$$

$$\frac{\partial}{\partial z}\left(\frac{fu}{T}\right) = +\frac{\partial}{\partial y}\left(\frac{g}{T}\right)$$

Completing the differentiation and rearranging we have:

$$\frac{\partial v}{\partial z} = \frac{g}{fT}\frac{\partial T}{\partial x} + \frac{v}{T}\frac{\partial T}{\partial z}$$

$$\frac{\partial u}{\partial z} = -\frac{g}{fT}\frac{\partial T}{\partial y} + \frac{u}{T}\frac{\partial T}{\partial z}$$

In the normal case, when the second term in each of the above equations can be neglected, these become the thermal wind equations:

$$\frac{\partial v}{\partial z} = \frac{g}{fT}\frac{\partial T}{\partial x}$$

$$\frac{\partial u}{\partial z} = -\frac{g}{fT}\frac{\partial T}{\partial y}$$

Here the primary difference between terrestrial and Martian velocities will be due to the smaller g and larger temperature gradients on that planet, which will act in opposite directions, so that again we can expect comparable winds to those on Earth.

We can now use the geostrophic and thermal wind equations to estimate the wind velocities to be expected on Mars. For a temperature of $250°$K and a latitude of $45°$ we know:

$$f = 10^{-5}$$
$$\rho = 3.7 \times 10^{-5}$$
$$g = 376$$
$$T = 250.$$

In order to apply the equations we must estimate the pressure gradients and the thermal gradients. Let us consider a one percent pressure change in 400 km. From Figure 4 we can estimate a temperature gradient of $30°$K over 3000 km. Putting these values into the equations we find geostrophic winds of 10^4 cm/sec, and thermal

winds over a vertical height of ten km of 10^4 cm/sec. These speeds, of the order of 100 miles per hour, are typical of those found in the Earth's atmosphere at high altitudes and are about an order of magnitude greater than speeds estimated from movements of yellow clouds and two orders of magnitude greater than are needed to explain the speed of the wave of darkening. It should be noted that we have neglected friction and surface turbulence in all these calculations, and our results apply only to winds at altitudes greater than about one km. Nearer the ground the winds will be slower and more variable.

This is as far as it seems desirable to continue investigation of Martian weather. The concepts we have developed will be useful in the following two sections where the evaporation of the polar caps and circulation of water vapor are discussed.

ENERGY BALANCE IN THE POLAR CAPS

In order to determine surface temperatures we have considered the planetary surface to be in radiative equilibrium with the solar flux. In so doing we have assumed that all the incident solar energy went into heating of the surface. If the surface is of a material which undergoes a phase change under the influence of solar radiation it is also necessary to account for the amount of energy absorbed in accomplishing the phase change. Thus for the poles, where the ice can undergo a phase change, we can write the energy balance equation as:

$$H(1-A) = \sigma T_o^4 + L\rho_i \Delta z$$

where H is the insolation, L is the energy required per gram to evaporate ice, ρ_i is the density of the ice, and Δz is the thickness of ice evaporated in unit time. Putting in numerical factors this equation becomes:

$$H(0.5) = 5.67 \times 10^{-5} T_o^4 + 2.82 \times 10^{10} \rho_i \Delta z$$

We know H; if we also knew T_o we could solve for $\rho_i \Delta z$ and so find the amount of water in the polar cap. However, we have seen that the observations are not nearly good enough to let us solve in this straightforward manner, nor are they likely to be in the near future.

G. de Vaucouleurs [1954] has discussed this problem and has stated that the probable thickness of the caps is a cm or so, although he does not indicate the line of reasoning which led him to this conclusion. He states that, while the actual thickness is known only to order of magnitude, that the profile (relative thickness as a function of latitude) can be determined with considerable accuracy. To do so requires accurate values of the relative amount of heat available to vaporize the caps, as a function of latitude, which is as uncertain as the absolute values. I feel that his profile is as uncertain, or more so, as his order-of-magnitude value for the thickness.

The most recent detailed discussion of the polar caps is
by Lebedinskii and Salova [1962]. In order to solve the equation
on the preceeding page these authors assume, in view of the
dryness of the Martian atmosphere, that the relative humidity
above the caps is zero. In this case the evaporation rate, and
rate of energy used in evaporation, are known functions of the
temperature: $\rho_i \Delta z = \rho_i \Delta z(T_o)$. With this assumption the equation
can be solved for T_o and $\rho_i \Delta z$ as a function of time and place, and
a summation with respect to time, from the beginning of spring
until the disappearance of the cap, gives the thickness of the
cap at any latitude. They obtain a thickness of about 0.01 gm cm^{-2},
for a total quantity of water of 2 X 10^{15} gm.

This is an ingenious method, and, in view of the reasonableness
of the results obtained, it is an attractive one. Unfortunately
it is incorrect, both conceptually and numerically.

To show this it is necessary to digress and find the equation
for the evaporation rate of ice as a function of temperature.
In view of the errors made in this respect by L. and S., it seems
advisable to sketch the derivation here, following that on
pp 742-744 of Estermann [1955]. We consider a solid in equilibrium
with its vapor; in this case the same number of molecules pass
across the interface in each direction per unit time and unit area.
We assume further that the rate of molecules leaving the solid
is independent of the presence of the vapor; this assumption
has been experimentally verified. Thus the problem is reduced
to finding the number of molecules per second which strike unit
area of the surface from the vapor in equilibrium.

Consider the contribution from an infinitesmal cone defined
by an angle θ from the normal to the surface. The velocity of the
molecules is v and their number density is N. We call the flux
from this elementary cone dF. The probability of the velocity
being directed toward the unit area considered is $1/4\pi$. The area
seen by the incoming particals is $\cos \theta$. The distance of the
unit volume from the unit area does not enter in, since the flux

from a unit volume varies as r^{-2}. Thus we can write:

$$dF = \frac{Nv}{4\pi} \; 2\pi \, \sin\theta \, \cos\theta \; d\theta$$

Now integrate over θ:

$$F = \int_{0}^{\pi/2} 1/2 \; Nv \, \sin\theta \; (d \, \sin\theta)$$

$$= Nv/4$$

Substituting for N and v from kinetic theory gives the flux in molecules per second as:

$$F = p \, (2\pi mkT)^{-1/2}$$

The final form for E, the evaporation rate in cgs units, is:

$$E = p \sqrt{\frac{\mu}{2\pi RT}} \quad gm \; cm^{-2} \; sec^{-1}$$

The formula used by L. and S. for the evaporation rate does not explicitly involve p, the vapor pressure, at all. They give no derivation or source for their formula.

The vapor pressure of water is experimentally known over the temperature range of interest and is given in Figure 5. The dependence of p on T dominates the variation of E with T. This variation, computed from the above formula and the data of Figure 5, is shown in Figure 6. It should be emphasized that these figures are for evaporation into a vacuum or gas of zero humidity; to find the evaporation rate in the case of non-zero humidity, β, these results must be multiplied by the factor $(1 - \beta)$.

To compare the results of Figure 5 with those given by the Russian authors let us consider the temperature of 220°K. The Russian evaporation rate at this temperature is 10^{-8}, while the value from Figure 5 is $10^{-3.4}$. At T = 260°K the corresponding numbers are 8×10^{-5} and $10^{-1.6}$. It is clear that, since the numbers used by L. and S. for the evaporation rate are off by

FIGURE 5

Vapor Pressure of H₂O

Log P (in dynes cm⁻²) as a function of temperature (in °K).

FIGURE 6

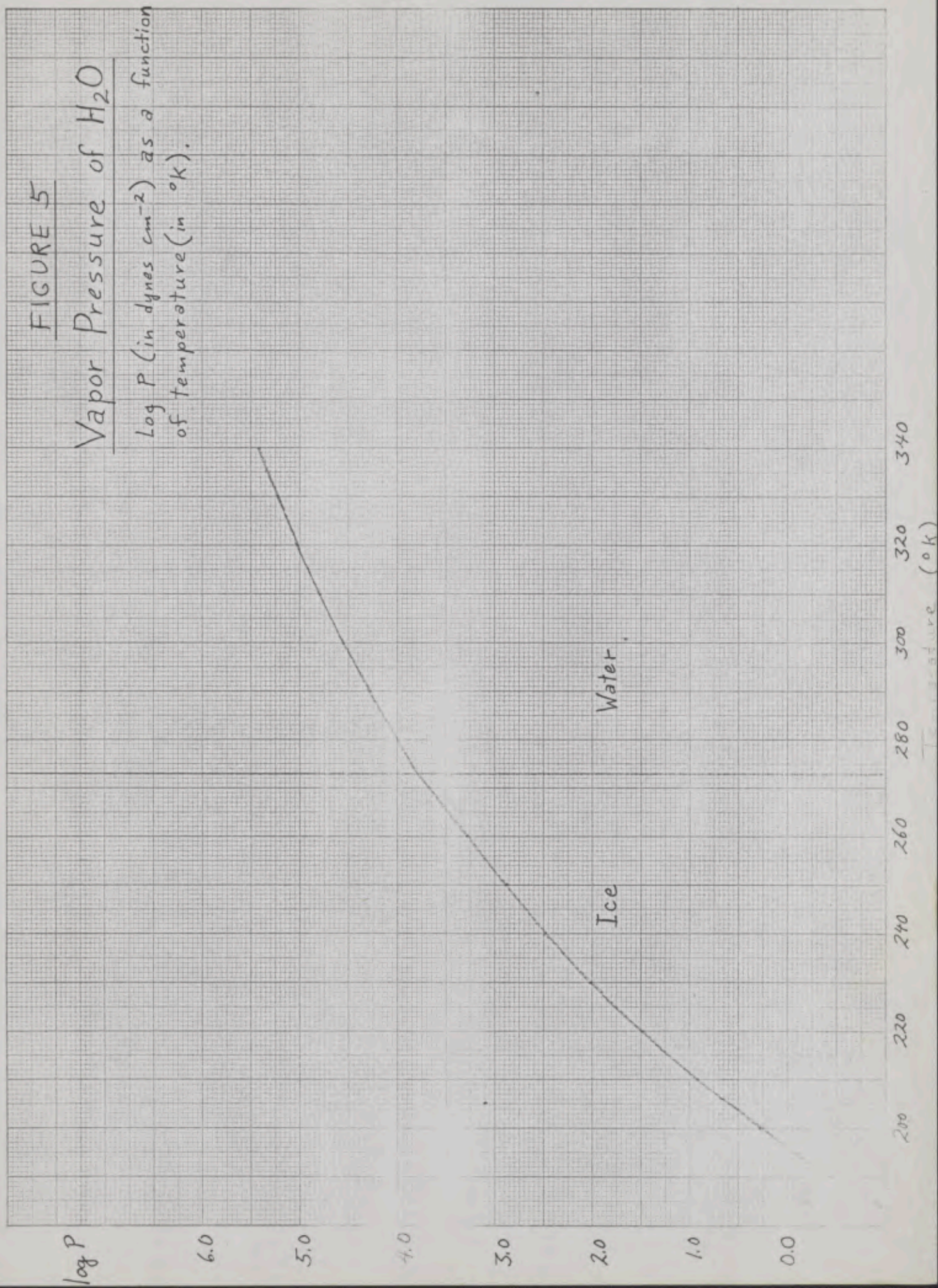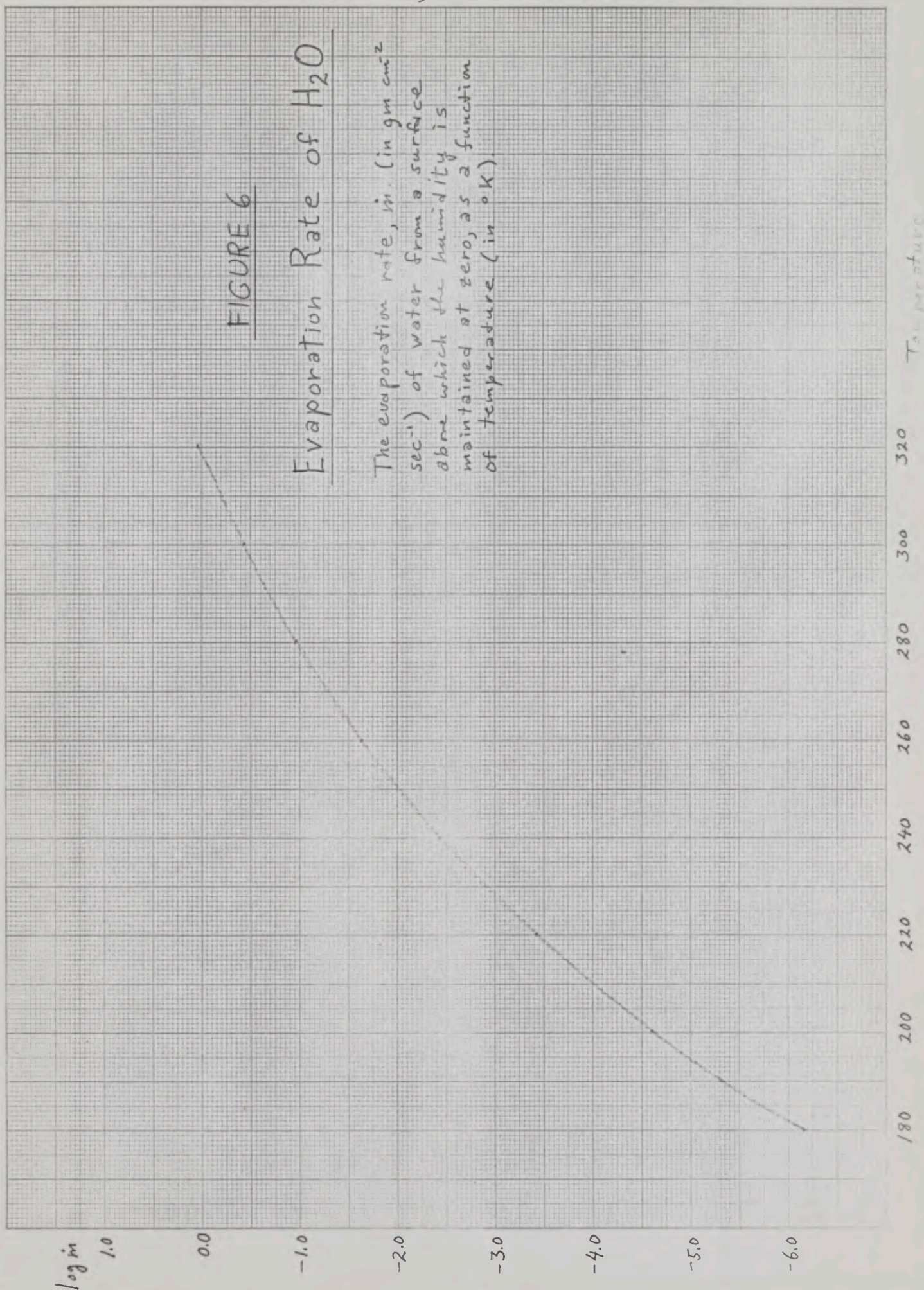## Evaporation Rate of H_2O

The evaporation rate, $\dot{m}$, (in gm cm^{-2} sec^{-1}) of water from a surface above which the humidity is maintained at zero, as a function of temperature (in °K).

log $\dot{m}$

1.0

0.0

-1.0

-2.0

-3.0

-4.0

-5.0

-6.0

180   200   220   240   260   280   300   320   Temperature

a factor of about 10^4, that they underestimate the thickness of
the caps by that amount. The correction of this error makes a
striking difference in their conclusions.

Conceptually, it seems clear that the assumption that the
humidity at the surface is zero is not a good one. Our terrestrial
experience is that evaporation rates are always controlled by the
rate at which vapor is transported from a thin saturated layer
in contact with the liquid surface: hence the efficacy of blowing
on hot soup. If evaporation rates were anything like those given
in Figure 5, we would have lakes and oceans vaporizing at rates
of meters per hour. It is obvious that they do not do this,
even when the air a few meters above the surface is far from
saturated. In order to apply the approach of L. and S. and determine
the thickness of the polar caps from the evaporation rate, we must
estimate the evaporation rate from atmospheric parameters, not
by ignoring the atmosphere.

If we correct the error of 10^4 made by the Russian authors
we can use their data to get a thickness of 100 gm cm^{-2} for
polar caps evaporating into zero humidity. This will surely
be an upper limit to the thickness of the caps. We can also
get an estimate of the thickness by comparing the insolation with
the temperatures observed by Gifford [1956] and assuming the
excess energy all went into vaporization of the ice. The results
are anywhere from 0.1 to 10 gm cm^{-2} depending on how the data are
interpreted, so that this is not a satisfactory method, even if
we take the data at face value.

An accurate computation of atmospheric transfer of water
vapor upward from the surface is not possible with present knowledge,
but it is possible to make some estimates. First let us consider
thermal diffusion. The vapor pressure at the surface is given by
the surface temperature, and we shall assume that the atmosphere
imediately above the surface is also at this temperature. The
diffusion equation is identical to the heat flow equation we used
to estimate the energy conducted through the planetary surface, and we

find an approximate solution in the same manner as before. The
equation is:

$$\frac{dm}{dt} = - D \frac{d\rho_v}{dz}$$

where dm/dt is the mass transported across unit area in unit time,
ρ_v is the vapor density at the surface, and D is the diffusion
constant, given by λv, where λ is the mean free path and v is
the average velocity of the molecules. The mean free path and
velocity are given by:

$$\lambda = \frac{kT}{p\sigma} \qquad v = \sqrt{\frac{3kT}{M}}$$

where σ is the collisional cross-section and M is the mass of
a molecule. This gives the functional form of D to be:

$$\frac{D}{D_0} = \left(\frac{T}{T_0}\right)^{3/2} \left(\frac{p_0}{p}\right)$$

From the Handbook of Chemistry and Physics we obtain the value
of D for water vapor diffusing into air at a pressure of one
atmosphere and a temperature of 281°K. This gives, for D, when
p is expressed in mb:

$$D = 0.239 \left(\frac{T}{281}\right)^{3/2} \left(\frac{1000}{p}\right)$$

The resulting values of D are plotted in Figure 7.

Often the diffusion rate is dominated, not by thermal diffusion,
but by turbulent diffusion. The eddy diffusion coefficient, D_e,
is computed in the same way as the thermal coefficient, except that
the mean free path is replaced by the mixing length and the
molecular velocity by the mean turbulent velocity. It can be
shown [Hess 1959, pp 283-287] that, under simplifying assumptions,
D_e is given by the product of the density, the von Karmann constant,
the height above the surface, and the friction velocity which is
related to the wind speed. D_e is known experimentally for the
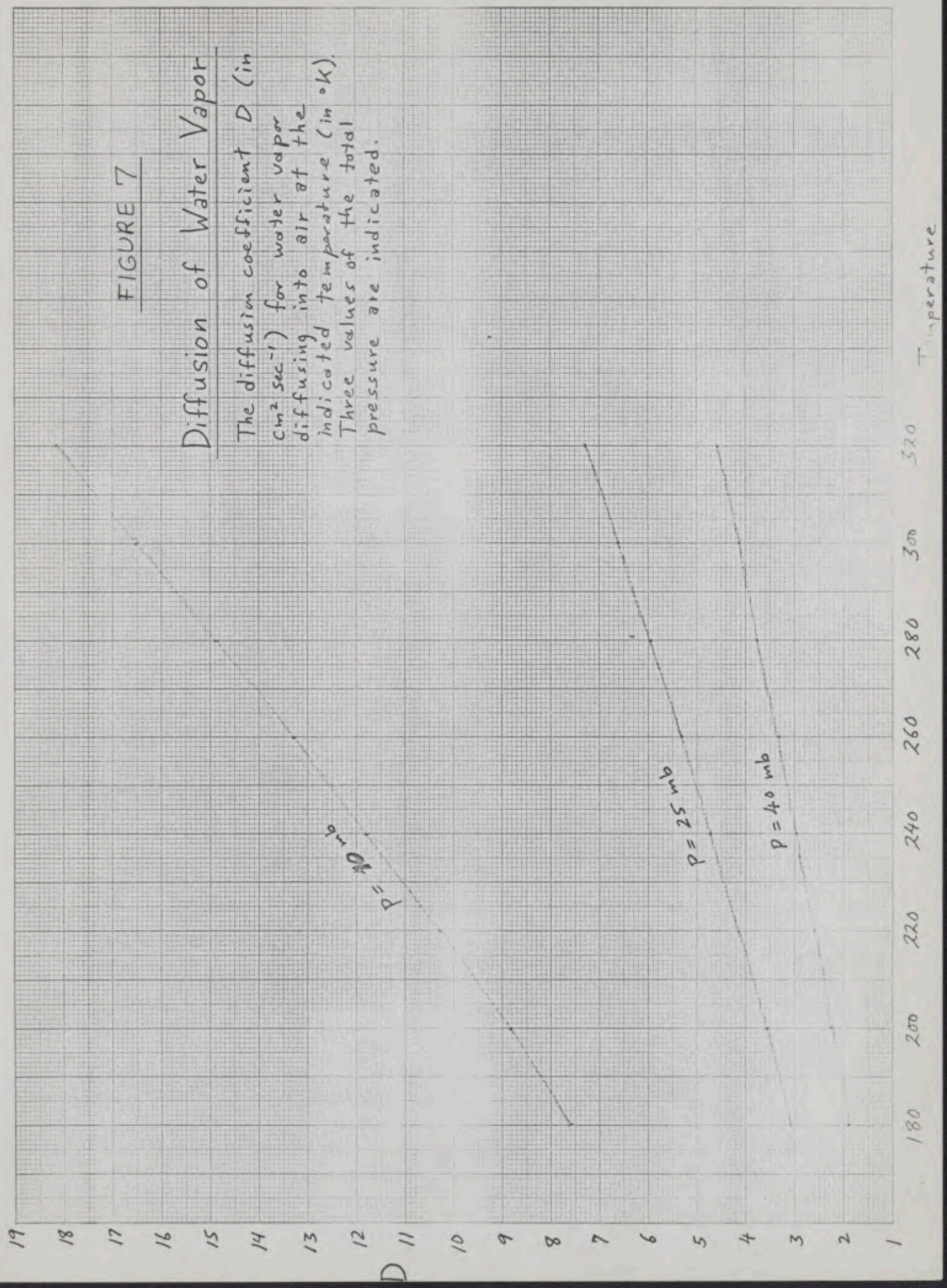Earth's atmosphere; from Lettau [1951], p 322, the values typical

FIGURE 7

Diffusion of Water Vapor

The diffusion coefficient D (in $cm^2 sec^{-1}$) for water vapor diffusing into air at the indicated temperature (in $^\circ K$). Three values of the total pressure are indicated.

near the surface of the Earth are 10^2 to 10^5 cm^2 sec^{-1}. The Martian winds are comperable to those on Earth, so the only scaling that we must do to the terrestrial value is due to the difference in density, which is a factor of 40. This indicates values for D_e no greater than the values of D indicated in Figure 7. It seems that we can use the values from that Figure in the expectation that they are good to order of magnitude.

To find the mass that diffuses from the surface we first find the height to which the material has diffused, given by the square root of D·t, and then put finite differences into the equation at the top of the preceeding page. The results, in mass evaporated per cm^2 in one hour, are given as a function of temperature in Figure 8, using just the values of D given in Figure 7.

From the collection of facts presented so far in this paper, I will now attempt to construct the events which take place in the Martian polar regions and to estimate the amount of water on the planet.

FIGURE 8

Thermal Diffusion Rate of Water Vapor

The log of the mass in $gm \, cm^{-2}$ which diffuses from the surface ice or water in one hour at the indicated temperature, for an atmospheric pressure of 25 mb.
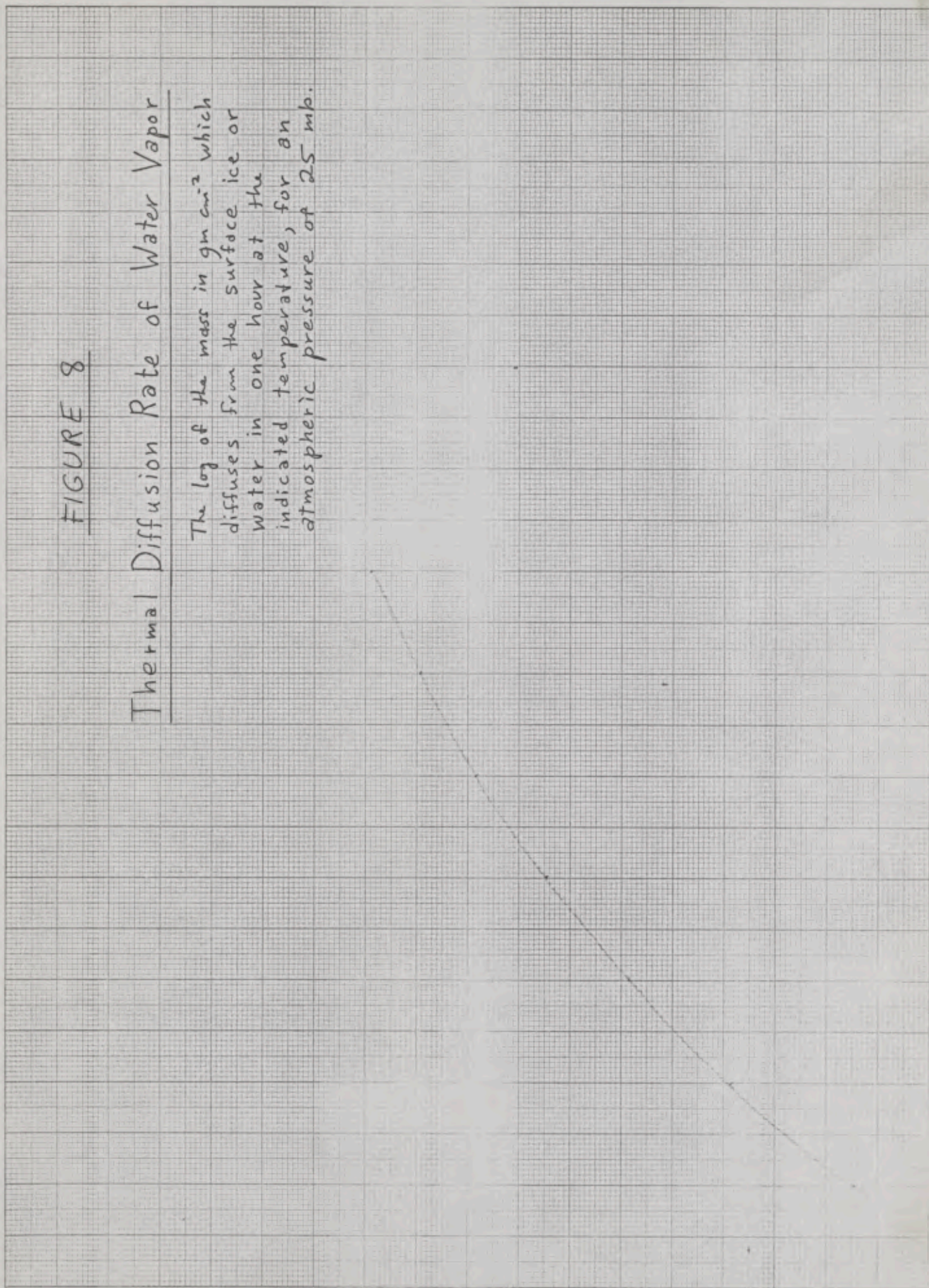
Temperature of surface

log m

THE POLAR CAPS: A SYNTHESIS

In discussing the seasonal changes which the polar caps undergo, we must not forget the rigorous diurnal variation in conditions which the Martian surface experiences. All events are dominated by the alternation of solar heating and rapid nighttime cooling. The one exception is the arctic regions near the solstices; here, in the "land of the midnight sun", the surface is subjected to continuous, if variable, solar radiation.

During the long winter night the polar cap forms under temperatures which never rise above 200°K. Under these conditions the surface vapor pressure is less than 10^{-3} mb and the evaporation is negligable. Only after the equinox does solar radiation begin to raise the temperature into the range of interest.

Consider a point at latitude -70°; the temperature will be somewhere between the two curves for that latitude given in Figure 1. Consider $\eta = 300^\circ$; the temperature will be about 240°K at noon. From Figure 5 the surface vapor pressure is 50 dyne cm^{-2}, giving a mixing ratio for water vapor of about 10^{-3}. The mass evaporated in one hour is, from Figure 8, about 10^{-3} gm; during the whole day this will still be less than 10^{-2} gm. At this rate the polar cap will vaporize in only a few days assuming it to be only a mm or so thick. But as night comes the temperature drops again to 200° or less, at which temperature the atmosphere is unable to support the water vapor; at 200° the saturation vapor pressure is only 10^{-3} mb, so that for a mixed layer at this temperature with a thickness of 1 km, no more than 10^{-4} gm cm^{-2} can remain in the atmosphere. The rest condenses back on to the surface. Part of the water vapor will, of course, be removed by advection, and the variations in wind pattern from one year to the next will produce the observed variations in the rate of shrinkage of the polar cap.

From Figure 3 we can see that the edge of the cap has a temperature near 260 degrees; this indicates about 10 times the evaporation rate computed above. But most of the time the

temperature is much below the noon value, so a reasonable
estimate of the vapor content of the atmosphere above the cap
during the day is 10^{-2} gm cm^{-2}. Let us now suppose that there is
a mean wind off the polar cap of 10^3 cm sec^{-1}, and take the
effective edge of the cap across which this wind blows to be
1000 km long. Then the mass of water blown off the cap in
a day is 10^{14} gm. The melting of the polar cap takes about
100 days, so this gives for the total amount of water in the
cap 10^{16} gm. It seems probable that this is correct to within
one order of magnitude. It gives a thickness of the polar cap,
if it has unit density, of 1 mm, and it predicts an average
water vapor content of the atmosphere when the polar cap has
evaporated of 10^{-2} gm cm^{-2}, assuming that there are no other
sources or sinks (such as living things) of water. When one of
the polar caps is present, we would expect to see an order of
magnitude less water vapor in the atmosphere, unless we were
able to look specifically over the poles, where there would be
more.

If there is 10^{-2} gm cm^{-2} of water vapor in the atmosphere,
and if we assume a constant mixing ratio throughout the troposphere
of thickness 10 km, then the mixing ratio is 3×10^{-4}. This means
condensation will occur at about 215°K. From these arguments,
we expect to see no clouds in the daytime, but clouds or frost
at night and at the cooler pole are expected. If most of the
water condenses out on or near the surface at night, this may be
the explaination of the "dawn haze" which is often observed.
Also, if such a mechanism exists to concentrate water near the
surface, this makes the water much more available to living things
than if it remained mixed throughout the troposphere.

Returning to the evaporation of the polar cap, I have pointed
out that during the night most of the water evaporated during the
daytime condenses back on to the surface. But as the spring
progresses more and more of the pole is exposed to continuous
sunlight. Not much will evaporate anyway until the temperature
nears 240°, which does not happen at the pole itself until a

couple of weeks before the solstice. Everything balances out so
that the regions which have continuous day are too cold to take
advantage of it. However, at η = 320 degrees the two effects
join forces: the edge of the cap (at latitude -70°) retreats
into the land of the midnight sun and the temperature of the
edge at noon *reaches* a maximum. The result is clearly shown
on Figures 2 and 4 in the increased rate of recession which takes
place at this time. Only at the solstice itself, and for a very
few weeks afterward, does the pole warm up enough to finally shed
its thin deposit of ice. During the northern summer the temperatures
are about 20° cooler than the conditions we have been discussing,
so that the maximum temperature of the pole is only about 225°K,
not enough to evaporate the ice cap there. The northern cap
shows strong recession at the latitude 70° point mentioned above,
but shortly thereafter the recession stops. It is interesting to
speculate as to how much water might be present in that perpetual
ice-cap; I see no reason why it could not be as large as all
the rest of the water put together.

Finally, the question must be discussed: is there any
liquid water on Mars? Water can exist in the liquid phase only
when the partial pressure of water is greater than the triple point
pressure and the temperature is above 0°C, but below the boiling
point, specified by the temperature at which the vapor pressure
exceeds the total pressure. For a surface temperature of 25 mb
the temperature range for liquid is from 273° to 295°; for 10 mb
it is from 273° to 280°. Over the bulk of the planet the surface
partial pressure never approaches 6.1 mb, the triple point pressure,
but we have seen that over the polar caps the atmosphere is essentially
saturated at the surface. The conclusion is that whenever the
edge of the polar cap has a temperature above 273°, liquid water
will be present there, although probably only in a narrow strip,
the width of which depends on local advection. It seems quite
possible that the dark collar around the retreating cap is due to
wetting of the ground by liquid water, although the extensive blue
lakes proposed by Lowell are not possible.

CONCLUSIONS

We have found that the polar caps are of water, that their thickness is of the order of 1 mm, and that the total amount of water is about 10^{16} gm, which corresponds to 10^{-2} gm cm^{-2} if distributed uniformly on the planet.

During this century many unsuccessful attempts have been made to observe water vapor spectroscopically on Mars; the best of these have put an upper limit near 10^{-3} gm cm^{-2} averaged over the planet. At the last opposition positive results were finally obtained by three groups of observers, and two of these have been reported in press. Dollfus [1963], observing during January 1963 at $\eta = 125$, reported 2×10^{-2} gm cm^{-2}, a surprisingly high result in terms of former failures. Kaplan, Münch, and Spinrad [1964], in April 1963 at $\eta = 165$, found $1.4 + .7 \times 10^{-3}$ gm cm^{-2}. These observations were both during the northern spring; one would expect more of the polar cap to have melted in April than in January and so to have seen more water then. While the observations certainly confirm the order of magnitude of the estimates we have made, they are puzzling in their lack of agreement with each other. More work is needed; for the next few years oppositions will be getting better, so we can expect much more and better data soon.

I think that it is important to observe at each quadrature for the next few years, since observations can not be made near opposition, so as to obtain an idea of the variation of atmospheric water vapor with Martian season. Then we can re-examine the conclusions of this paper and perhaps finally determine how much water there is on Mars.

REFERENCES

de Vaucouleurs, G. [1954] Physics of the Planet Mars, Faber
and Faber, London.

Dollfus, A. [1957] Ann. d'Astrophys. Supp. No. 6.

Dollfus, A. [1961] in Planets and Satellites, ed. Kuiper and

Dollfus, A. [1963] Middlehurst, U. of Chicago Press, Chicago.
Académie des Sciences, Séance du 1er Avril, 1963, p 1009

Estermann, I. [1955] in Thermodynamics and Physics of Matter, Vol. 1,
ed. F. D. Rossini, Princeton U. Press, Princeton.

Fleagle, R. G., and Businger, J. A. [1963] An Introduction to
Atmospheric Physics, Academic Press, New York.

Gifford, F. [1956] Ap. J. 123, 154.

Goody, R. M. [1957] Weather 12, 3.

Hess, S. L. [1959] Introduction to Theoretical Meteorology,
Holt, New York.

Kaplan, L. D., Münch, G., and Spinrad, H. [1964] Ap. J. 139, 1.

Kellogg, W. W., and Sagan, C. [1962] The Atmospheres of Mars
and Venus, National Academy of Sciences
Publication 944.

Kiess, C. C., Karrer, S., and Kiess, H. K. [1960] P.A.S.P. 72, 256.

Kuiper, G. P. [1952] in The Atmospheres of the Earth and Planets,
ed. Kuiper, U. of Chicago Press, Chicago.

Kuiper, G. P., and Middlehurst, B. M. [1961] editors, Planets and
Satellites, U. of Chicago Press, Chicago.

Lebedinskii, A. I, and Salova, G. I. [1962] Soviet Astronomy - AJ
6, 390. (English translation)

Lettau, H. [1951] in Compendium of Meteorology, ed. T. F.
Malone, Waverly Press, Baltimore.

Lowell, P. [1906] Mars and Its Canals, Macmillan, London.

Mintz, Y. [1961] in The Atmospheres of the Earth and Planets, by
Kellogg and Sagan, N. A. S. Publication 944.

Ohring, G., Brooks, E. M., Marino, J. [1964] The Meteorology of
Mars and Venus, GCA Tech. Report No. 64-4-N.

Pettit, E. [1961] in Planets and Satellites, ed. Kuiper and
Middlehurst, U. of Chicago Press, Chicago.

Sagan, C. [1964] paper to be given at Flagstaff AAS meeting.

Slipher, E. C. [1962] Mars, Sky Publishing Corp., Cambridge.